Impact of Weddell Sea deep convection on natural and anthropogenic carbon in a climate model

Raffaele Bernardello¹, Irina Marinov¹, Jaime B. Palter², Eric D. Galbraith³, and Jorge L. Sarmiento⁴

¹Department of Earth and Environmental Science, University of Pennsylvania, Philadelphia, Pennsylvania, USA,
²Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada,
³Department of Earth and Planetary Sciences, McGill University, Montreal, Quebec, Canada,
⁴Atmospheric and Oceanic Sciences Program, Princeton University, Princeton, New Jersey, USA

Abstract

A climate model is used to investigate the influence of Weddell Sea open ocean deep convection on anthropogenic and natural carbon uptake for the period 1860–2100. In a three-member ensemble climate change simulation, convection ceases on average by year 1981, weakening the net oceanic cumulative uptake of atmospheric CO₂ by year 2100 (−4.3 Pg C) relative to an ocean that has continued convection. This net weakening results from a decrease in anthropogenic carbon uptake (−10.1 Pg C), partly offset by an increase in natural carbon storage (+5.8 Pg C). Despite representing only 4% of its area, the Weddell Sea is responsible for 22% of the Southern Ocean decrease in total climate-driven carbon uptake and 52% of the decrease in the anthropogenic component of oceanic uptake. Although this is a model-specific result, it illustrates the potential of deep convection to produce an intermodel spread in future projections of ocean carbon uptake.

1. Introduction

The high-latitude Southern Ocean (SO) is thought to play a pivotal role in the preindustrial air-sea partitioning of carbon because of its ability to connect the atmosphere to the vast volume of the deep ocean through the formation of Antarctic Bottom Water (AABW) [Sarmiento and Toggweiler, 1984; Sarmiento and Orr, 1991; Marinov et al., 2006; Séférian et al., 2012]. However, discrepancies among different techniques to estimate ocean (C$_{o}$) content are especially pronounced for deep waters of southern origin, making the role of AABW uncertain [Vázquez-Rodríguez et al., 2009].

AABW presently forms at specific locations on the Antarctic continental shelf [Orsi et al., 1999], while, in the past, this water mass was also known to form during open ocean deep convection events in the Weddell Sea (WS) [Martinson et al., 1981; Gordon, 1982; Killworth, 1983]. The open ocean convective pathway of AABW was well-documented in satellite observations which revealed an ice-free, convective region in the WS with a size of 2–3 × 10⁵ km² for three consecutive winters in the 1970s [Casey, 1980]. The absence of this pathway in the last 40 years coincides with observed freshening and stratification of the SO [de Lavergne et al., 2014], and warming and volume loss from the AABW layer, possibly linked with a slowdown in its formation rate [Purkey and Johnson, 2012].

Meanwhile, open ocean convection in the WS is a prominent feature of many climate models under preindustrial forcing [Heuze et al., 2013]. It is important to note that the coarse resolution hampers the correct representation of shelf processes in global models and therefore may place too much importance on deep open ocean convection. Nonetheless, deep convection has been shown to weaken and cease in warming simulations [de Lavergne et al., 2014], raising the specter of a warming-induced perturbation to ocean carbon uptake and storage. These models vary significantly in the strength and duration of open ocean convection and its sensitivity to warming. Importantly, the timing of the slowdown may give rise to intermodel spread in the evolution of the modeled air-sea exchange of CO₂. Although previous studies have used a water mass framework to characterize the SO carbon uptake [e.g., Séférian et al., 2012], the specific role of subpolar SO deep convection on air-sea CO₂ exchange in models has yet to be explored.

Here we use a coupled climate model to understand the role of WS convection and its disappearance on the ocean’s storage of natural and anthropogenic carbon. Because our model simulates WS convection events under preindustrial conditions [Galbraith et al., 2011] and their abrupt cessation under warming...
[Bernardello et al., 2014; de Lavergne et al., 2014], it provides an ideal tool for exploring these potential feedbacks to the simulated carbon cycle.

The response of ocean carbon uptake to changes in deep ocean convection is complex. During Weddell Sea convection AABW is formed through heat loss from Circumpolar Deep Water (CDW) exposed to the atmosphere. CDW has, on average, been isolated from the surface of the ocean for millennia [DeVries and Primeau, 2011]. During this time, CDW accumulates high dissolved inorganic carbon (DIC) concentrations as a consequence of the remineralization of sinking particulate organic matter. Because of these high DIC concentrations, a WS convection event would be expected to produce a transient rise in atmospheric CO₂ via natural carbon (Cₙᵃᵗ) outgassing from newly exposed, supersaturated CDW prior to the anthropogenic addition of CO₂ to the atmosphere. This transient rise in atmospheric CO₂ would be counterbalanced partly by the reestablishment of an equilibrium between the atmosphere and surface mixed layer and partly by the accumulation of remineralized carbon in the deep global ocean, leading to a steady state on timescales longer than those of convection. Thus, the suppression of convection-driven outgassing under anthropogenic warming might represent a negative feedback on atmospheric CO₂ by allowing the ocean to retain an increasing load of remineralized DIC. However, there is also a positive feedback that occurs from reduced oceanic uptake of Cₙᵃᵗ in a nonconvecting ocean, as will be described below. Hence, the expected net cumulative effect of the suppression of open ocean deep convection on global ocean CO₂ uptake is far from obvious.

After describing the model and our simulation design (section 2), we synthesize the impact of convection on Cₙᵃᵗ storage in the WS under preindustrial conditions (section 3) and the implications of a climate-driven convective shutdown on air-sea CO₂ fluxes (section 4). Finally, we offer conclusions and explore additional questions left for future study in section 5.

2. Model and Simulation Design

All simulations were carried out using CM2MC [Galbraith et al., 2011], the Geophysical Fluid Dynamics Laboratory Climate Model version 2 with Modular Ocean Model version 4p1 at coarse resolution. CM2Mc uses the same physical oceanic and atmospheric code as the Earth System Model with the Modular Ocean Model (ESM2M) [Dunne et al., 2012], with minor alterations as required to adjust to the coarser discretization. The CM2Mc ocean uses a tripolar grid with nominal 3° resolution and 28 vertical levels with a finer resolution toward the surface. Eddy mixing is parameterized following Gent and McWilliams [1990] with a spatially varying diffusion coefficient [Griffies et al., 2005]. Ocean biogeochemistry is solved by the Biogeochemistry with Light, Iron, Nutrients, and Gases model, with a representation of carbon chemistry [Galbraith et al., 2010; Bernardello et al., 2014].

The model was initialized as described in Galbraith et al. [2011] and spun-up for 4500 years. In order to account for sensitivity to initial conditions, we extracted atmospheric and oceanic conditions at three points during the last 1000 years of the spin-up: immediately before, during, and after a convection period. These three states were used to initialize three independent simulations, producing an ensemble, under each of the following four scenarios:

A. Warming_increasingCO₂: A climate change simulation using prescribed atmospheric concentrations for greenhouse gases and aerosols recommended by the Coupled Model Intercomparison Project Phase 5 (CMIP5). We use the historical forcing (1860–2005) and the Representative Concentration Pathway 8.5 (2006–2100), detailed in Meinshausen et al. [2011]. Ozone, including contemporary depletion and its predicted future recovery, is also prescribed [Gionni et al., 2011].

B. Warming_constantCO₂: In this simulation, the climate responds to the radiative forcing due to historical+RCP8.5, but the ocean chemistry is not affected by the rise in atmospheric CO₂.

C. Control_increasingCO₂: A preindustrial climate simulation, where greenhouse gases, aerosols, and ozone are kept at the value of year 1860. However, ocean chemistry is affected by the increase in atmospheric CO₂ according to Historical+RCP8.5, allowing Cₙᵃᵗ to invade the ocean through air-sea gas exchange.

D. Control_constantCO₂: A preindustrial control simulation, in which radiative forcing is at preindustrial level (as in Control_increasingCO₂) and ocean chemistry responds to a constant preindustrial atmospheric CO₂ concentration (as in Warming_constantCO₂).
Area of study

Figure 1. Yearly averaged Weddell Sea convection cycles in a 500 year segment of preindustrial simulation. (a) Area of study; (b) area-integrated $CO_2$ flux and total mass of DIC inside the study area; averages of the study area for (c) sea ice concentration and salinity (average 0–50 m depth); (d) water temperature and mixed layer depth.

By comparing the air-sea $CO_2$ fluxes ($\phi$), integrated over the area of study (Figure 1a), in these simulations, we can quantify the direct impact of the suppression of WS convection on the ocean's ability to retain $C_{nat}$ and absorb $C_{ant}$. The difference between Warming_increasing$CO_2$ and Control_increasing$CO_2$ ($A - C$) reveals the influence of WS convection (among other climate-driven changes in the area) on the total carbon ($C_{nat} + C_{ant}$) ocean uptake ($\Delta \phi_{Total}$), while Warming_constant$CO_2$ minus Control_constant$CO_2$ ($B - D$) reveals the impact on the $C_{nat}$ flux only ($\Delta \phi_{Natural}$). Thus, the difference ($(A - C) - (B - D)$) reveals the impact of climate change on the $C_{ant}$ uptake ($\Delta \phi_{Total} - \Delta \phi_{Natural} = \Delta \phi_{Anthropogenic}$). For the last 1000 years of the spin-up we saved yearly averaged outputs to investigate the characteristics of regular convective cycles. Monthly averages were saved for simulations A, B, C, and D. Since open ocean convection occurs in winter, we defined a convective year as any year in which the September mixed layer was deeper than 2000 m over any portion of the study area in simulations A, B, C, and D. By comparing the September mixed layer with its annual average, we found that annually averaged mixed layer for nonconvective years was never deeper than 60 m. Therefore, during the 1000 year segment, we used a mixed layer depth threshold of 60 m to distinguish convective from nonconvective years.

CM2Mc response to historical+RCP8.5 forcing is discussed in Bernardello et al. [2014] while de Lavergne et al. [2014] compare CM2Mc to CMIP5 models and to observations for aspects related to SO deep convection. For the present study we have also calculated mean CM2Mc water masses characteristics following the procedure described by Sallée et al. [2013] as a further model evaluation (see supporting information).

3. Effects of the Simulated Weddell Polynya on Climate and Natural Carbon Storage

Under preindustrial conditions, deep convection in the Weddell Sea is characterized by a regular periodicity in CM2Mc, as illustrated by the vertical homogenization of temperature from the surface to 4000 m (Figure 1d). Non-convective periods occur, on average, every 60 years and last for approximately 20 years. During these non-convective periods, warm and salty water originating from CDW is transported into the WS at depth and accumulates between 200 m and 3000 m, creating a temperature inversion that is stabilized by the halocline. Reduced vertical mixing between the surface and this warmer and saltier layer contributes to cooling and freshening at the surface and coincides with an expansion of sea ice (Figure 1c). Because the surface freshening dominates in setting the evolution of the pycnocline during non-convective
periods, density stratification increases in spite of the subsurface warming. In our model, WS convection is periodically triggered by changes in the precipitation-evaporation balance, which lead to a low-frequency modulation of surface salinity [Galbraith et al., 2011]. Once convection begins, the upward flux of heat from the subsurface ocean prevents the formation of winter sea ice, and a large polynya develops, qualitatively similar to the polynya observed in the 1970s, though often significantly larger (up to 8 times, Figure 2). Its position migrates within the trapezoidal area depicted in Figure 1a, which defines our area of study.

During convective periods, CM2Mc global surface air temperatures rise by 0.4°C relative to nonconvective periods, linked to the heat released from the deep ocean (not shown). A similar response was described for another model that simulates WS convection events with centennial scale variability [Latif et al., 2013]. Furthermore, by exchanging with the vast Cnat pool of the deep ocean, each WS convective period promotes an average CO2 outgassing of about 6.1 Pg C. Over the 60 years of a convective period, this amount of CO2 corresponds to about 0.1 Pg C yr\(^{-1}\) (Figure 1b) and contributes a considerable fraction of the total variability in the preindustrial control simulation: the standard deviation of the globally integrated annual flux of CO2 in Control_constantCO2 is just 0.16 Pg C yr\(^{-1}\). Following the heat release, the erosion of the temperature inversion promotes the stabilization of the water column and the reestablishment of a surface halocline. This allows the transition to a nonconvective period, the expansion of sea ice, the steady subsurface accumulation of heat and carbon, and the beginning of a new cycle.

4. Collapse of the Weddell Polynya and Response of the Air-Sea CO2 Fluxes

In our simulations, surface freshening, resulting from an increased precipitation minus evaporation balance over most of the SO, suppresses deep convection in the WS under historical+RCP8.5 forcing (Figure 3). The last convective year for each of the three ensemble members is 1952, 1987, and 2005. Following this cessation, heat continuously accumulates at depth, but the surface freshening provides increasing stability and prevents the onset of convection until the end of the simulations in 2100. Under preindustrial
atmospheric CO₂ concentrations, simulated WS deep convective events cause outgassing of Cnat (Figure 4a, Control_constantCO₂) because the pCO₂ at the sea surface during these events is higher than its atmospheric counterpart, as expected.

However, in addition to its impact on Cnat, the suppression of convection in the WS impacts the uptake and storage of anthropogenic CO₂ from the atmosphere. If open ocean convection persists in spite of rising atmospheric CO₂, as is simulated in Control_increasingCO₂, the CO₂ outgassing during convective events gradually weakens. This weakening begins as soon as anthropogenic emissions start to rise, due to the reduction in the air-sea CO₂ gradient during convective events. When atmospheric pCO₂ climbs to the pCO₂ of the exposed deep waters, the sign of the total flux reaches zero, after which convective events become net ingassing events (Figure 2). Essentially, the pCO₂ of water exposed during deep convection is set by the deep ocean carbon reservoir, which increases very slowly compared to atmospheric pCO₂. Because the atmospheric pCO₂ rises while the pCO₂ of exposed deep waters stays relatively constant, this part of the ocean becomes a sink of anthropogenic CO₂, relative to the preindustrial baseline. The very same deep mixing that released carbon under preindustrial pCO₂ then becomes a conduit that transfers absorbed carbon to depth. This redistribution allows for the continued uptake of anthropogenic CO₂ throughout the duration of the convection event. Thus, anthropogenic CO₂ uptake in the WS is 46% lower when convection is suppressed (last two rows of Table 1).

Additional outgassing regions are visible in the September 1950 snapshot (Figure 2), mainly between 60⁰E and 120⁰E in the Indian Ocean sector of the SO. The southernmost area, between 60⁰S and 48⁰S, lies within the region of Ekman-driven upwelling where CO₂-rich CDW is transported into the surface mixed layer. Between 48⁰S and 36⁰S, the mixed layer deepens along the path of

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**Figure 3.** DIC concentration averaged over the study area for the three ensemble members of the Warming_increasingCO₂ simulation. September mixed layer is superimposed (black line). Precipitation-evaporation flux and salinity (average 0–50 m depth) for the same area are also shown for each ensemble member.

**Figure 4.** (a) Air-sea CO₂ flux integrated over the area of study for the four simulations. (b) Cumulative climate-driven changes in CO₂ fluxes, calculated from the ensemble averages as: \( \Delta \Phi_{\text{Natural}} = \text{Warming\_constantCO₂} - \text{Control\_constantCO₂} \) \( \text{(B} - \text{D)} \), \( \Delta \Phi_{\text{Total}} = \text{Warming\_increasingCO₂} - \text{Control\_increasingCO₂} \) \( \text{(A} - \text{C)} \), \( \Delta \Phi_{\text{Anthropogenic}} = \Delta \Phi_{\text{Total}} - \Delta \Phi_{\text{Natural}} \). In both panels thick lines refer to the ensemble mean while thin lines show ±1 standard deviation relative to the three-member ensemble.
considering the wide spectrum of intensity and frequency of SO deep convection in CMIP5 models and has remained shut since the 1970s in nature [de Lavergne et al., 2014].

Considering the wide spectrum of intensity and frequency of SO deep convection in CMIP5 models and the range of their response to climate change [Heuze et al., 2013; de Lavergne et al., 2014], a similarly wide
spectrum in total carbon uptake and $C_{\text{ant}}$ interior distribution is to be expected. Roughly a third of models have no SO deep convection during their preindustrial simulations; of the two thirds that do convect, the strength of preindustrial convection and the sensitivity to anthropogenic climate change vary widely [de Lavergne et al., 2014]. The longer simulated deep convection persists in the face of warming, the more $C_{\text{ant}}$ is taken up in the SO. Similarly, models with larger areas of convection expose more of their interior ocean volume to high atmospheric pCO$_2$ and store a higher fraction of the absorbed $C_{\text{ant}}$ in AABW. We suggest that variability in SO deep convection could be a major source of intermodel spread for the carbon cycle. Although our results are based on one coarse-resolution model only, our model is near the multi-model average of CMIP5 simulations with respect to the percentage of convective years (66% for CM2Mc and 63% for CMIP5 average) and the size of the convective region during the preindustrial (7.6 x 10$^5$ km$^2$ for CM2Mc and 9.3 x 10$^5$ km$^2$ for CMIP5 average) [de Lavergne et al., 2014], suggesting the impact on the carbon cycle may be fairly representative of the convecting models. However, convection collapses early in the period of historical forcing relative to many of the models, which we would expect to lead to large weakening of the oceanic $C_{\text{ant}}$ sink.

In summary, we have shown that deep convection in the polar SO has a strong impact on the global ocean carbon uptake in transient climate simulations with a climate model. Considering that deep convection in the polar SO is a common and widely variable feature among the latest generation of climate models we stress here the necessity to better understand its role in determining the intermodel spread in ocean carbon uptake. The development of proxy records for past variability in Southern Ocean deep convection would allow better characterization of this process in the real world.

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